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## RESEARCH ARTICLE

10.1002/2015JF003517

## Special Section:

Glacier Surging and Ice Streaming

## Key Points:

- Thermal switches can explain the behavior of small glaciers during LIA climate cycle
- The surges of large glaciers cannot be explained by thermal switching
- Enthalpy cycling can explain dynamic switches of both small and large Svalbard glaciers

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## Thermal structure of Svalbard glaciers and implications for thermal switch models of glacier surging

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**Abstract** Switches between cold- and warm-based conditions have long been invoked to explain surges of High Arctic glaciers. Here we compile existing and new data on the thermal regime of six glaciers in Svalbard to test the applicability of thermal switch models. Two of the large glaciers of our sample are water terminating while one is land terminating. All three have a well-known surge history. They have a thick basal layer of temperate ice, superimposed by cold ice. A cold terminus forms during quiescence but is mechanically removed by calving on tidewater glaciers. The other three glaciers are relatively small and are either entirely cold or have a diminishing warm core. All three bear evidence of former warm-based thermal regimes and, in two cases, surge-like behavior during the Little Ice Age. In Svalbard, therefore, three types of glaciers have switched from slow to fast flow: (1) small glaciers that underwent thermal cycles during and following the Little Ice Age (switches between cold- and warm-based conditions), (2) large terrestrial glaciers which remain warm based throughout the entire surge cycle but develop cold termini during quiescence, and (3) large tidewater glaciers that remain warm based throughout the surge cycle. Our results demonstrate that thermal switching cannot explain the surges of large glaciers in Svalbard. We apply the concept of enthalpy cycling to the spectrum of surge and surge-like behavior displayed by these glaciers and demonstrate that all Svalbard surge-type glaciers can be understood within a single conceptual framework.

## 1. Introduction

The idea that glacier surges can be explained by changes in basal thermal conditions has a long history [e.g., Clarke, 1976; Mayer *et al.*, 2011; Robin, 1955, 1969]. Simply stated, glaciers where the bed temperature is below the pressure melting point (hereafter, cold-based glaciers) may be susceptible to thermal and dynamic instabilities if ice thickening raises basal temperatures and driving stresses, leading to increased rates of ice deformation and frictional heating. Positive feedbacks between strain heating and ice creep may raise the ice temperature to the pressure melting point, whereupon continued heat generation will melt ice and produce water at the bed. If more meltwater is created that can be discharged, basal water pressure rises, reducing basal drag and accelerating glacier sliding. This accelerated motion (surge) yields geometric changes contributing to stretching and thinning of the glacier. Drainage of basal water and conductive cooling return the glacier to its initial state, whereupon the cycle can begin once more. It has long been recognized that the existence of temperate surging glaciers means that thermal instability cannot provide a general explanation of all surges [Clarke, 1976; Clarke and Goodman, 1975; Jarvis and Clarke, 1975], although thermal instability remains a viable surge mechanism for glaciers with basal temperatures near the melting point.

Inspired by observations of Bakaninbreen, Svalbard [Murray *et al.*, 2000], and Trapridge Glacier, Yukon Territory, Canada [e.g., Clarke *et al.*, 1984; Fowler *et al.* [2001] developed a system of equations representing the coupled thermodynamics, hydrology, and dynamics of thermally controlled surging for a glacier underlain by deformable till and permeable bedrock. They considered three states of the bed; “warm” (till unfrozen, with the freezing front at the ice-till interface), “cold” (freezing front beneath the till-bedrock interface), and “subtemperate” (till partially frozen), varying in extent in space and time. Simulations with a simplified 1-D lumped parameter version of the model exhibited a range of oscillatory behaviors, including “fast” and “slow” surges. A key finding was that oscillatory behavior occurs when the hydraulic conductivity of the bed is small and that a steady state occurs where water can escape from the system fast enough to evacuate basal

meltwater. The existence of a frozen glacier tongue and subglacial and proglacial permafrost may play a key role in controlling water flux, and consequently patterns of surge propagation and termination [e.g., Murray *et al.*, 2000; Smith *et al.*, 2002].

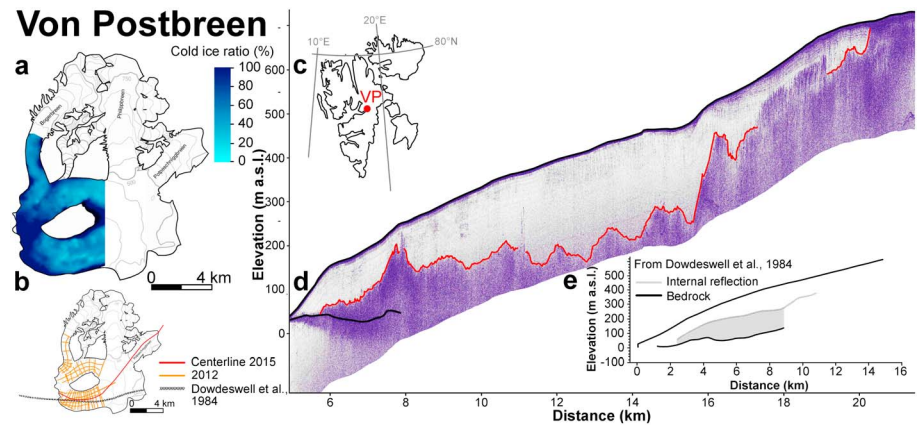
The aim of this paper is to test the hypothesis that thermal switching can explain surging behavior of polythermal glaciers in Svalbard. A wide range of glaciers are known or inferred to have surged in the archipelago, including small valley glaciers, tidewater glaciers, and land-terminating outlet glaciers, with varying thermal regimes from cold based to polythermal [Dowdeswell *et al.*, 1984; Hagen *et al.*, 1993; Jiskoot *et al.*, 2000; Lovell *et al.*, 2015; Sund *et al.*, 2009]. This variability makes Svalbard an ideal location to assess the role that switches in basal thermal regime play in the initiation and evolution of surges. In this study, we use ground-penetrating radar (GPR) data to investigate the thermal regime of six glaciers in Svalbard, and other lines of evidence to infer their longer-term thermal evolution and its relationship to their dynamic history. Published GPR data are available for many Svalbard glaciers, including most of the glaciers in this study [e.g., Bamber, 1987; Björnsson *et al.*, 1996; Dowdeswell *et al.*, 1984], although these data were acquired and processed using a wide variety of methods. In this study, therefore, we collected new GPR profiles to obtain a homogeneous data set for a sample of glaciers with a range of sizes, terminus types, and dynamic histories. We show that thermal switches are implicated in some surges but cannot explain others. We go on to consider whether multiple mechanisms must be invoked to explain Svalbard surges or if all can be understood within a single conceptual framework.

## 2. Study Sites

The Svalbard archipelago, Arctic Norway, contains one of the highest proportions of surge-type glaciers in the world [Sevestre and Benn, 2015], with the percentage of glaciers classified as surge-type varying from 13% to 90% [Jiskoot *et al.*, 2000; Hagen *et al.*, 1993]. To some extent, this range of estimates reflects different definitions of “surges,” and the criteria employed for their identification, an issue we return to later in this paper. Six glaciers in the main island of Spitsbergen were investigated in this study, chosen to provide a representative sample of different glacier sizes and geometries. These are the following: Von Postbreen, a large land-terminating glacier; Kongsvegen and Tunabreen, large tidewater glaciers; and Midtre Lovénbreen, Tellbreen, and Longyearbreen, small valley glaciers. Previous studies have compared GPR profiles with borehole temperature data from two of the glaciers (Kongsvegen and Midtre Lovénbreen [Björnsson *et al.*, 1996; Hagen *et al.*, 1993]). Inclusion of these glaciers in our sample thus provides important elements for interpreting GPR data from other sites.

Von Postbreen (78.45°N, 17.82°E) is the largest glacier in our sample with a length of 31 km and an area of 168 km<sup>2</sup> [König *et al.*, 2013] (Figure 1c). The glacier has two tongues; both of which now terminate on land. Von Postbreen is known to have surged in 1870 [De Geer, 1910], when it advanced into Tempelfjorden and deposited a large moraine [Forwick *et al.*, 2010]. Von Postbreen was formerly confluent with the tidewater glacier Tunabreen, which has an independent surge history (see below). In addition, Von Postbreen's northern tributary Bogebreen surged in 1980 [Dowdeswell *et al.*, 1984]. Von Postbreen is currently quiescent. Numerous studies have investigated the thermal structure and bed topography of Von Postbreen, including gravity surveys [Oélsner, 1966] and echo sounding surveys [Dowdeswell *et al.*, 1984]. The latter revealed the two-layered thermal structure of the glacier with cold layer overlying thick temperate ice (Figure 1e).

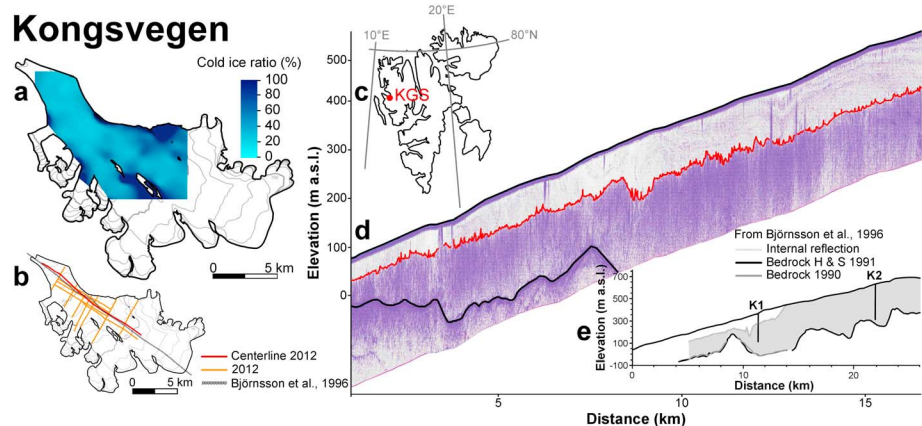
Kongsvegen (78.79°N, 13.15°E) is 24 km in length and 108 km<sup>2</sup> in area [König *et al.*, 2013] (Figure 2c). It terminates at the head of Kongsfjorden and is partially grounded and partially calving at its front. It is confluent with Kronebreen, one of the fastest flowing glaciers of the archipelago [Lefauconnier *et al.*, 1994]. Historical records indicate three large advances of the Kongsvegen-Kronebreen system: in 1800, 1869, and circa 1948, suggesting a cycle of 70–80 years [Lefauconnier, 1987; Liestøl, 1988; Voigt, 1965; Woodward *et al.*, 2002]. The surge of 1948 resulted in an advance of 2 km and is fully attributable to Kongsvegen, which compressed and deflected the tongue of Kronebreen. The glacier is currently in quiescence, with velocities of only 1.4 to 3.6 m a<sup>-1</sup> [Melvold and Hagen, 1998]. Since its last surge, the glacier has been retreating at a rate of 250 m a<sup>-1</sup> [Lefauconnier *et al.*, 1994]. Elevation changes on the glacier indicate that it may be building up to a new surge. Between 1966 and 2005, Kongsvegen thickened by about 0.5 m a<sup>-1</sup> in its upper parts and thinned by ~1 m a<sup>-1</sup> at lower elevations [Nuth *et al.*, 2010]. Published work shows that the glacier consists of temperate ice overlain by a cold surface layer [Björnsson *et al.*, 1996]. The cold layer becomes thinner with increasing elevation, and the glacier is temperate over its entire depth in its upper reaches. Direct temperature



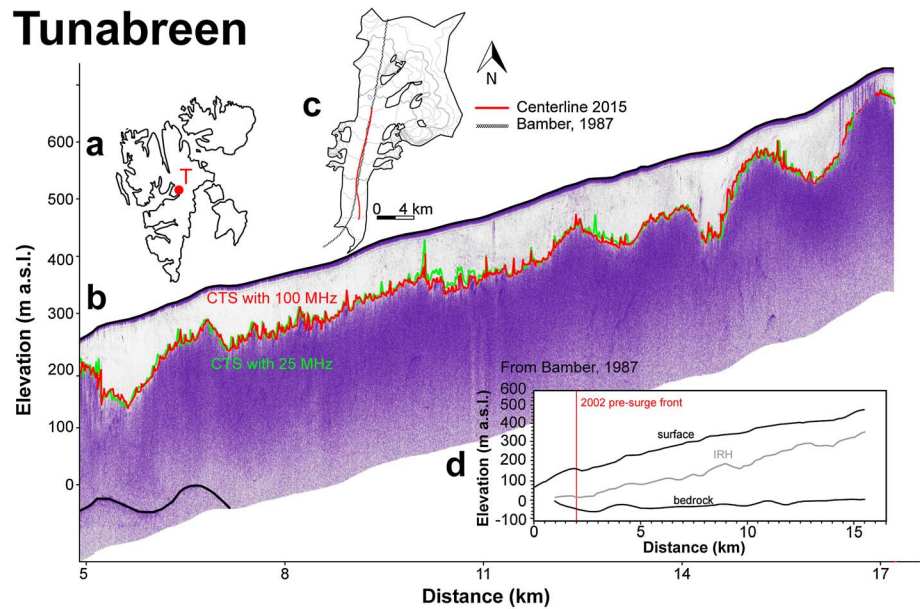
**Figure 1.** Compilation of new and existing GPR data collected on Von Postbreen. (a) Ratio of cold ice thickness over the glacier thickness, in percentage. (b) Location of the 2012 and 2015 survey lines and of the centerline profile collected in 1980 [Dowdeswell *et al.*, 1984]. (c) Location of Von Postbreen (VP) in Svalbard. (d) Centerline radargram collected in 2015, corrected for elevation. The top pick marks the glacier surface, while the bottom pick follows the bed reflector. The red line follows the cold-temperate transition surface (CTS). (e) Interpretation of the centerline radargram collected by Dowdeswell *et al.* [1984]. Modified from Dowdeswell *et al.* [1984].

measurements in two boreholes [Hagen *et al.*, 1993] showed ~150 m of warm ice below a ~100 m thick cold layer in the ablation zone (375 m above sea level (asl)) and >300 m of warm ice beneath only 15 m of cold ice in the accumulation zone (625 m asl) (Figure 2e).

Tunabreen (78.56°N, 17.61°E) is a large tidewater glacier flowing from Filchnerfonna and Lomonosovfonna into Tempelfjorden (Figure 3c). The glacier covers an area of 163 km<sup>2</sup> and is 27.5 km long [König *et al.*, 2013]. Like Kongsvegen, Tunabreen is among the few glaciers of Svalbard known to have surged three times: first in the 1930s, then in the early 1970s, and between 2003 and 2005 [Fleming *et al.*, 2013; Flink *et al.*, 2015; Forwick *et al.*, 2010; Hodgkins and Dowdeswell, 1994] with each surge yielding an advance less extensive than the previous one. During its last surge, the glacier advanced 1.4 km into Tempelfjorden. The surges of Tunabreen are characteristically short. The 1970 surge is believed to have ended in 1971 [Flink, 2013; Plassen *et al.*, 2004], while the most recent surge lasted about 2 years. Tunabreen is separated from Von Postbreen by an ice-cored moraine deposited by the latter during its 1870 surge. The thermal regime of the glacier was determined in 1983 by the Scott Polar Institute during an extensive campaign collecting

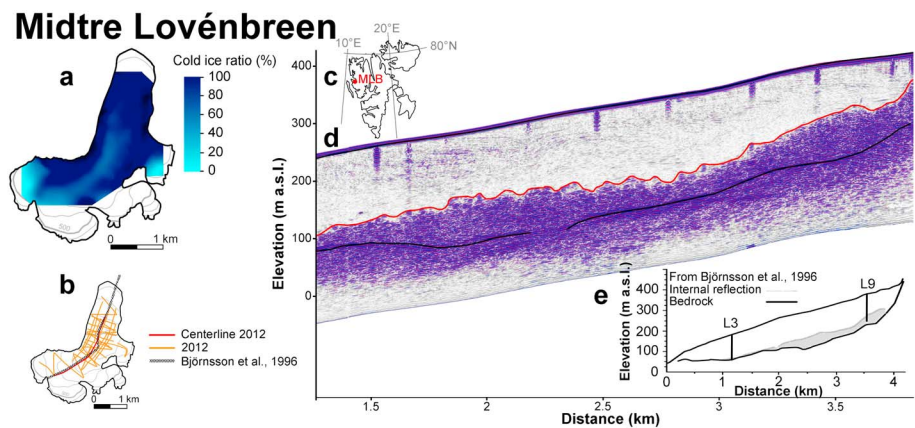


**Figure 2.** Compilation of new and existing GPR data collected on Kongsvegen. (a) Ratio of cold ice thickness over the glacier thickness, in percentage. (b) Location of the 2012 survey lines and of the centerline profile collected in 1990 and 1991 [Björnsson *et al.*, 1996]. (c) Location of Kongsvegen (KGS) in Svalbard. (d) Centerline radargram collected in 2012, corrected for elevation. The top pick marks the glacier surface, while the bottom pick follows the bed reflector. The red line follows the CTS. (e) Interpretation of the centerline radargram collected by Björnsson *et al.* [1996]. Modified from Björnsson *et al.* [1996]. K1 and K2 refer to two boreholes where temperature measurements were taken.

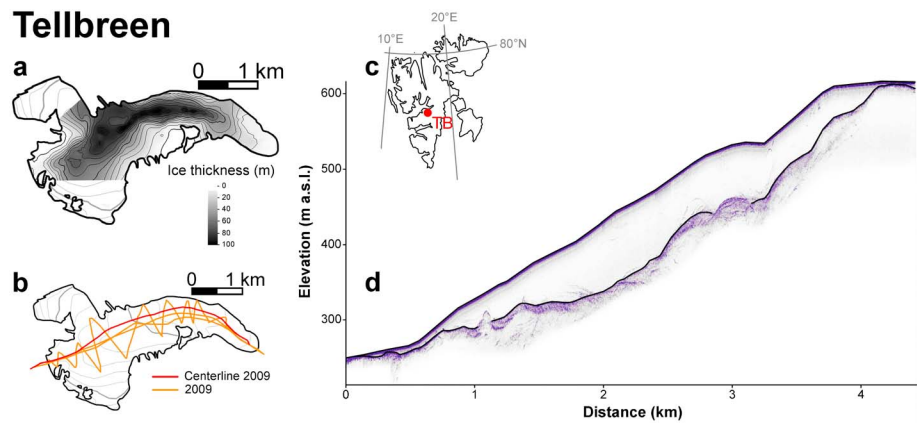


**Figure 3.** Compilation of new and existing GPR data collected on Tunabreen. (a) Location of Tunabreen (T) in Svalbard. (b) Centerline radargram of Tunabreen collected in 2015 with a 100 MHz antenna. The top pick marks the glacier surface, while the bottom pick follows the bed reflector. The red line follows the CTS picked on the data collected with the 100 MHz antenna, while the green line follows the CTS picked on data collected with the 25 MHz antenna along the same survey line. (c) Location of Tunabreen (T) in Svalbard. (d) Interpretation of the centerline radargram collected by *Bamber* [1987]. Modified from *Bamber* [1987]. The IRH is represented in grey. The position of the 2002 presurge front is indicated in red.

echo sounding data on 34 tidewater glaciers in Svalbard [Drewry and Liestøl, 1985]. A centerline profile collected at 60 MHz is available in *Bamber* [1987] (Figure 3d). Although it extends to the glacier terminus, neither the bed nor internal reflecting horizon (IRH) could be identified in this area due to multiple reflections from surface crevasses. Temperate ice was detected 2 km from the terminus and thickens with the glacier over 14 km. A surface layer of cold ice decreases in thickness from about 150 m to 100 m toward the upper end of the profile. The timing of the survey (1983) is particularly useful for the purposes of our study, because Tunabreen surged in 1970 and again in 2003–2005. The survey thus provides valuable insight into the presurge state of the glacier. The IRH identified by *Bamber* [1987] intersects the 2002



**Figure 4.** Compilation of new and existing GPR data collected on Midtre Lovénbreen. (a) Ratio of cold ice thickness over the glacier thickness, in percentage. (b) Location of the 2012 survey lines and of the centerline profile collected in 1990 [Björnsson et al., 1996]. (c) Location of Midtre Lovénbreen (MLB) in Svalbard. (d) Centerline radargram collected in 2012, corrected for elevation. The top pick marks the glacier surface, while the bottom pick follows the bed reflector. The red line follows the CTS. (e) Interpretation of a centerline radargram collected in by *Björnsson et al.* [1996]. Modified from *Björnsson et al.* [1996]. L3 and L9 refer to the location of two boreholes where temperature measurements were taken.

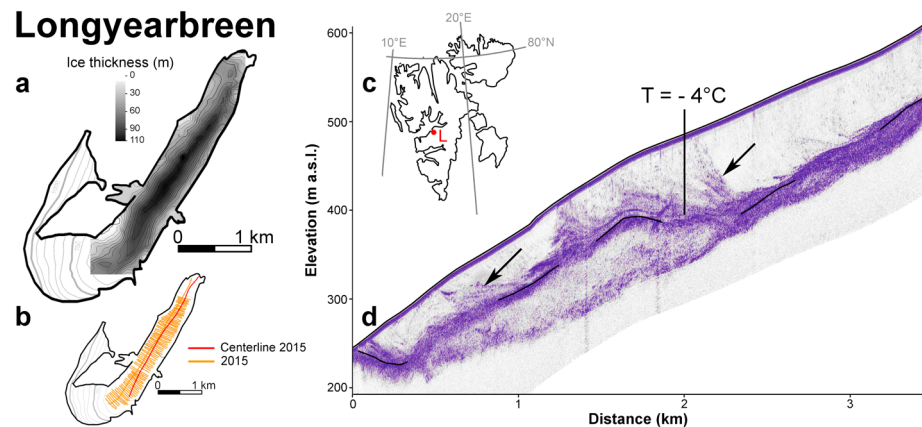


**Figure 5.** Compilation of new GPR data collected on Tellbreen. (a) Interpolation of the glacier thickness between the radar lines. (b) Location of the 2009 survey lines. (c) Location of Tellbreen (TB) in Svalbard. (d) Centerline radargram collected in 2009, corrected for elevation. The top pick marks the glacier surface, while the bottom pick follows the bed reflector.

calving front approximately 70 m above the bed. We believe that the position of the IRH is unlikely to have greatly changed in the intervening time, as heat conduction to the surface would be too slow to bring the front to a cold state. Thus, at surge onset, the bed of Tunabreen appears to have been entirely temperate due to efficient removal of ice by calving.

Midtre Lovénbreen (78.88°N, 12.03°E) is one of the most studied glaciers in Svalbard. It is about 4.4 km long, has an area of 5.2 km<sup>2</sup>, and consists of four accumulation basins [König *et al.*, 2013] (Figure 4c). Since the late nineteenth century, the glacier has retreated ~1 km [Glasser and Hambrey, 2001]. Centerline velocities range from 7.3 to 3.7 m a<sup>-1</sup> in the upper and lower tongue, respectively [Rippin *et al.*, 2005]. The glacier has the longest record of mass balance of the Arctic, covering nearly 47 years since 1967. It is currently thinning at an overall mean rate of 0.37 m a<sup>-1</sup> [Kohler *et al.*, 2007]. Published data [Björnsson *et al.*, 1996] indicate that the glacier is entirely cold where ice thicknesses are less than ~125 m and that the tongue is frozen to the bed (Figure 4e). Temperature data from two boreholes confirmed the depth of the cold-warm transition indicated by the GPR data [Björnsson *et al.*, 1996]. Glasser and Hambrey [2001] and Hambrey *et al.* [2005] used a wide range of evidence to demonstrate that Midtre Lovénbreen was more dynamically active in the late nineteenth century than now. The glacier was at or near its Neoglacial maximum in 1892, and photographs by Hamberg [1894] show a near-vertical terminal cliff traversed by subhorizontal debris bands, interpreted by Hambrey *et al.* [2005] as actively forming thrusts containing basal debris. Aerial photographs from 1948, 1966, and 1995 reveal networks of arcuate transverse fracture traces and debris bands progressively melting out of the glacier tongue, interpreted as evidence of formerly widespread thrust faulting. Hambrey *et al.* [2005] concluded that at the Neoglacial maximum, much of the lower glacier was actively sliding over soft, wet till, and that longitudinal compressive flow in the terminal zone resulted in thrust faulting and elevation of basal debris to the surface. Following this active phase, a major reduction in dynamic activity took place in the early twentieth century, concurrent with loss of mass and retreat. As the ice thinned, the winter cold wave was able to penetrate the bed and freeze the substrate, effectively ending subglacial deposition in the terminal region.

Tellbreen (78.25°N, 16.17°E) is a land-terminating valley glacier 4.5 km in length and 3.9 km<sup>2</sup> in area [König *et al.*, 2013] (Figure 5c), which flows from two cirques into a narrow tongue. Since the end of the LIA, the glacier has lost about 60 to 70% of its volume [Bælum and Benn, 2011]. Mass balance and flow velocities of Tellbreen have been measured since 2010 by students from the University Centre of Svalbard. Average surface velocity is ~1 m a<sup>-1</sup>, and mass balance is negative (annual balance of  $-2.3 \times 10^6$  m<sup>3</sup> w. eq. averaged over 2010–2014 (report AGF212 [2014])). Bælum and Benn [2011] reported results of GPR surveys that indicated the glacier was entirely or almost entirely cold, although they suggested that a thin basal layer of temperate ice may exist under the thickest part. Tellbreen has never been defined as a surge-type glacier although new evidence suggests otherwise. Lovell *et al.* [2015] analyzed structures exposed on the glacier surface and investigated englacial and subglacial ice exposures accessed via ice caves. The glacier is currently



**Figure 6.** Compilation of new GPR data collected on Longyearbreen. (a) Interpolation of the glacier thickness between the radar lines. (b) Location of the 2015 survey lines. (c) Location of Longyearbreen (L) in Svalbard. (d) Centerline radargram collected in 2015 with a 100 MHz antenna, corrected for elevation. The top pick marks the glacier surface, while the bottom pick follows the bed reflector. Arrows point to the englacial reflectors discussed in section 5.

frozen to its bed, although the ice-bed interface consists of subglacial traction till overlain by foliated, debris-rich basal ice. This indicates former wet-based conditions, with shearing of the basal till and stress-induced metamorphism of the basal ice. Above the basal ice layers, the englacial ice is traversed by dense networks of fracture traces, and arcuate bands of subglacially derived debris crop out on the glacier surface in the terminal zone. In combination, these features record a former phase of dynamic behavior, with a temperate bed, shearing of till and ice, and widespread crevassing and thrust faulting. This dynamic phase is inferred to have occurred at the LIA maximum, when the glacier was 60–70% larger, and approaching 200 m in thickness [Bælum and Benn, 2011].

Longyearbreen (78.17°N, 15.46°E) is a valley glacier 4.5 km long and 2.9 km<sup>2</sup> in area [König *et al.*, 2013] (Figure 6c). Although it is similar in length to Midtre Lovénbreen and Tellbreen, it is narrower with a single accumulation basin. Longyearbreen is currently a typical low-activity glacier, with a wide ice-cored moraine covering the terminus. GPR surveys have revealed a well-preserved V-shaped subglacial valley profile beneath the glacier [Etzel Müller *et al.*, 2000]. These surveys indicated that the glacier was entirely cold, with the possible exception of restricted areas of temperate ice close to the head of the glacier. An undisturbed in situ paleosoil and vegetation were found at the base of the glacier near its western margin 2 km upstream of the present front, dated from 1104 cal years B.P. [Humlum *et al.*, 2005]. This implies that this part of the glacier has been cold based since it advanced over this site at circa 900 A.D. Surface velocities are currently 1–4 m a<sup>-1</sup> [Etzel Müller *et al.*, 2000]. A mass balance program running from 1977 to 1992 showed that Longyearbreen was losing mass, with an average mean specific mass balance of  $-0.55 \text{ m a}^{-1}$  [Hagen *et al.*, 2003]. Studies have consistently classified Longyearbreen as nonsurge type [Etzel Müller *et al.*, 2000; Humlum *et al.*, 2005].

### 3. Methods

For this study, a total of 364 km of GPR profiles were collected on the six glaciers between 2009 and 2015. GPR has been widely used to map the thermal regime of Svalbard glaciers [Dowdeswell *et al.*, 1984; Hodgkins *et al.*, 1999]. Cold and pure ice is relatively transparent to electromagnetic waves, whereas temperate ice creates scatter at the result of diffractions from water-filled voids [Bamber, 1987]. The transition between cold, dry ice and temperate ice containing a small percentage of water yields strong scattering and hence can be easily detected [Navarro and Eisen, 2009]. Comparisons between borehole temperature measurements and ground-penetrating radar in Svalbard have shown that a continuous IRH corresponds to the cold-temperate transition surface (CTS) [Björnsson *et al.*, 1996; Ødegard *et al.*, 1997].

We employed a Malå radar system composed of a Malå XV monitor, a ProEx Professional Explorer control unit, and fixed unshielded “rough terrain” antennae, with elements oriented in-line. The system uses an impulse

**Table 1.** Surveys Details and GPR Settings Chosen for Each Survey<sup>a</sup>

Glacier Name	Survey Date	Antenna Used	Distance Tx-Rx (m)	GPS System	Acquisition Window (ns)	Sampling Frequency (MHz)	Stacks	Time Interval (s)
Von Postbreen	03/2012	100 MHz	2.2	Garmin eTrex	3106	621	32	0.18
Centerline	04/2015	100 MHz	2.2	Trimble SPS855	3243	574	16	0.1
Centerline	04/2015	25 MHz	6.2	Trimble SPS855	3237	397	4	0.1
Kongsvegen	04/2012	25 MHz	6.2	Garmin eTrex	3130	647	32	0.1
Tunabreen	04/2015	100 MHz	2.2	Trimble SPS855	3243	574	64	0.1
	04/2015	25 MHz	6.2	Trimble SPS855	3237	397	4	0.1
Midtre Lovénbreen	05/2012	25 MHz	6.2	Garmin eTrex	3130	647	32	0.1
Longyearbreen	04/2015	100 MHz	2.2	Trimble SPS855	2245	352	16	0.1
	04/2013	50 MHz	4.2	Trimble SPS855	2431	701	1	0.1
Tellbreen	03/2009	100 MHz	2.2	Garmin eTrex	2381	501	8	0.025

<sup>a</sup>Tx-Rx stands for transmitter, receiver. Time interval denotes the length of time between each transmitted pulse.

transmitter with a pulse repetition frequency of 100 kHz. The number of samples/traces goes from 128 to 2048, while the maximum number of stacks can reach 32,768. The sampling frequency can be adjusted between 0.4 and 100 GHz. All surveys were performed in autostacking mode to allow more flexibility in the data collection on site. Different center frequencies were used for different glaciers, ranging from 25 to 100 MHz (see Table 1). Although we expected the different antennae to have different sensitivities to dielectric discontinuities in the ice (see Table 2) we observed that the depth of the IRH was closely similar for the 25 and 100 MHz antennae on Von Postbreen and Tunabreen. Higher frequencies (>345 MHz [Pettersson, 2005]) are typically more sensitive to water content in the ice; therefore, traces from the 25 and 100 MHz antennae will tend to give us a minimum thickness for temperate ice. Due to limitations of the Malå system, measurement capabilities were restricted to 267 m of ice. For thicker ice, the depth of the bed could not be determined although the thermal structure of shallower ice could still be observed.

The system was towed behind a snowscooter driven at speeds below 20 km h<sup>-1</sup>, which equates to traces with a maximum spacing of 0.5 m when the signal is triggered every 0.1 s (see Table 1). All data were collected before the start of the melt season which would promote surface-antenna coupling that remains as constant as possible [Copland and Sharp, 2000]. No corrections were made for the presence of snow/firn at the surface of the glacier.

Due to the configuration of the antennae, only common offset surveys were performed. A constant velocity of 167 m μs<sup>-1</sup> was assumed for the propagation of radar waves in glacier ice for all glaciers [Paterson, 1994; Annan, 1999]. For horizontal and vertical positioning of the traces, two different GPS systems were used: a Garmin eTrex Legend and a Trimble SPS855 GPS (see Table 1). Uncertainties related to positioning are described below. GPS coordinates were recorded every 5–25 traces by the GPR system.

Data were processed using ReflexW version 7.5 (Sandmeier Scientific Software). The same processing sequence was used for bed and CTS mapping. Raw data were preprocessed in order to correct errors in the positioning from the GPS data, reset the time to zero when the signal was first sent, and remove repeated traces (see Table 3). The main processing consisted of Dewow to suppress low-frequency energy emitted by the field near the transmitter, then exponential gain function to increase the strength of the signal with depth. The dewow filter acts on each trace independently. A running mean value is calculated for each value of each trace and is then subtracted from the central point (ReflexW User's guide). A time window of 10 ns is used for the radargrams collected with the 100 MHz antenna, while a time window of 40 ns was employed for the radargrams collected with a 25 MHz antenna. Dynamic correction and interpolation of the traces to a regular spacing aim to prepare the data for the migration. They, respectively, look at individual traces and

**Table 2.** GPR Antennae and Resolutions

Antenna Frequency (MHz)	100	50	25
Wavelength (m)	1.67	3.34	6.68
Vertical resolution (m)	0.56 to 0.84	1.11 to 1.67	2.23 to 3.34
Lateral resolution at 100 m (m)	18.3	25.9	36.7
Lateral resolution at 200 m (m)	25.9	36.6	51.8



Preprocessing	Check coordinates, elevation Time zero correction Remove repeated traces
Processing	Dewow Dynamic correction Equidistant traces Kirchhoff migration Bandpass Butterworth Exponential gain function Time to depth conversion Topographic correction Picking

correct for the effects of antenna offset, and reposition the data with an equal distance between all traces. A Kirchhoff migration was then applied with a constant velocity of  $167 \text{ m } \mu\text{s}^{-1}$ . Kirchhoff migration is done in the  $x-t$  range and is applied on profiles where traces are equidistant. Migration aims to reset the position of the elements in the radargrams so that the reflections and diffractions move back to their true position; however, a 2-D Kirchhoff migration cannot correct off-transect reflectors that typically affect valley-parallel transects close to the valley sides. As migration increases noise, we filtered the unwanted high and low frequencies with a

band-pass butterworth filter. This sets the frequency spectrum below and above a “filter band” to zero (for 25 MHz: 0–80 MHz, for 50 MHz: 0–120 MHz, and for 100 MHz: 25–180 MHz). An exponential gain is applied next, in order to strengthen the signal with depth. The final steps consist of applying a time-to-depth conversion and correcting for the topography of the glacier surface. The effects of the main processing steps on the radar data are displayed in Figure 7. In the shown example (Von Postbreen), reflections from the glacier bed and the CTS can be easily differentiated on the lower 4 km of the glacier. The bed appears as a sharp, strong linear, and continuous reflector, while temperate ice is clearly distinguished from transparent and feature-free cold ice by numerous stacks of small hyperbolas.

To confirm our interpretations, we used published work by Björnsson *et al.* [1996], Hagen *et al.* [1993], Dowdeswell *et al.* [1984], and Humlum *et al.* [2005] to ground truth our data and compare the depth of the different reflectors to their observations, and direct temperature measurements. The retreat of the glaciers

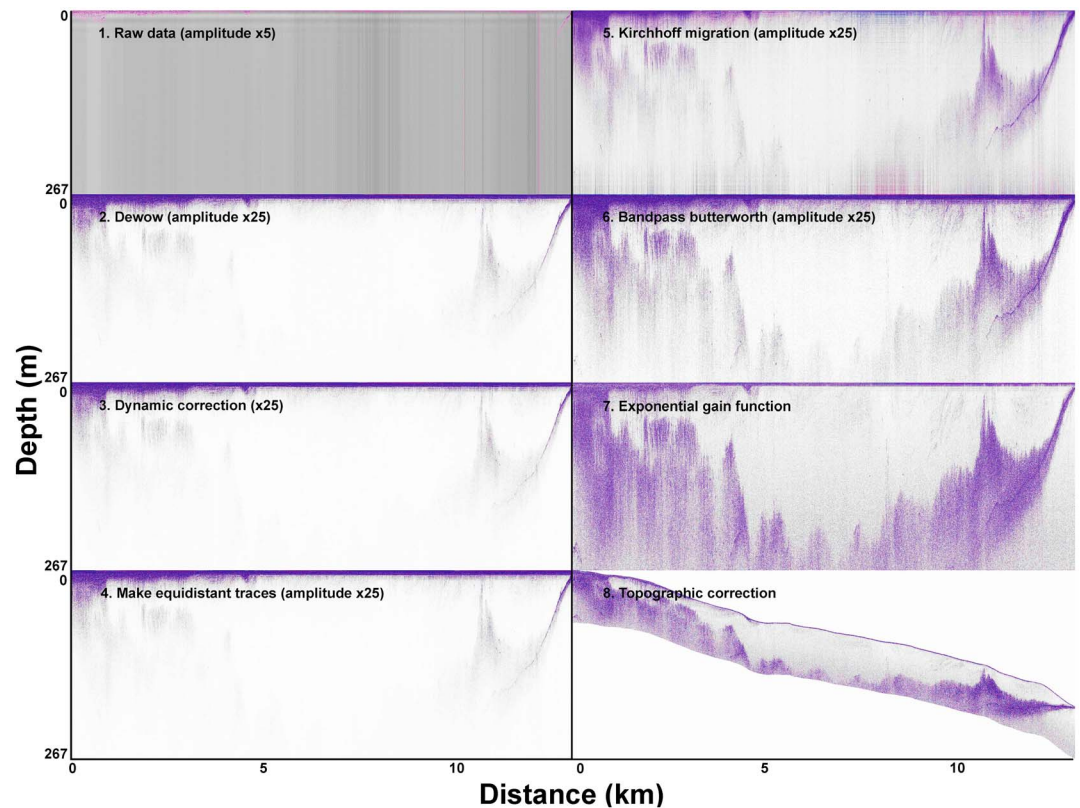


Figure 7. Main processing steps applied on the centerline profile of Von Postbreen.

**Table 4.** GPR Data Processing-Related Errors

	Mean Crossover Error (m)	Standard Deviation Crossover Error (m)	Mean Interpolation Error (m)	Standard Deviation Interpolation Error (m)
Von Postbreen	3.01	1.9	2.10	3.16
Kongsvegen	2.9	1.6	0.3	9.72
Midtre Lovénbreen	2.83	1.44	0.2	3.54
Tellbreen	1.47	0.55	0.53	3.76
Longyearbreen	2.7	1.03	1.59	3.56

between the old and new surveys was measured. The  $x = 0$  of the old profiles served as a reference for the new profiles, making a direct comparison possible. Although the surface topography and IRH depth have changed between the old and new surveys, the original findings were used as a guide to our interpretation. The glacier bed, surface, and CTS were then picked manually on the migrated profiles.

In order to improve the visualization of the spatial distribution of the thermal structure and bed topography of the glaciers, we use the software Petrel (Schlumberger) to interpolate the picked reflectors between the radar lines and extrapolate three nodes (150 m) beyond them. A convergent interpolation algorithm was used to interpolate between the lines. It is a general purpose algorithm which computes a surface (resolution  $50 \times 50$  m) through the picked lines distributed across the glaciers. Three different surfaces (interpolated horizons) were created, namely, glacier surface, CTS, and glacier bed. Glacier thickness was obtained by subtracting the bed depth to the glacier surface depth, while temperate ice thickness was obtained by subtracting the bed depth to the CTS depth.

### 3.1. Uncertainties

Uncertainties in our results related to the resolution of our technique are described in Tables 1 and 3. The vertical resolution of our radar system is estimated as being equal to one half or one third of the wavelength of the antenna frequency [Annan, 1992], giving a resolution of 0.56 to 0.84 m for the 100 MHz antenna and to 2.23 to 3.34 m for the 25 MHz antenna, assuming a wave velocity of  $167 \text{ m } \mu\text{s}^{-1}$ . Horizontal resolution is defined as the diameter  $D$  of the first Fresnel zone, as is estimated by Fowler [1990] as equal to:

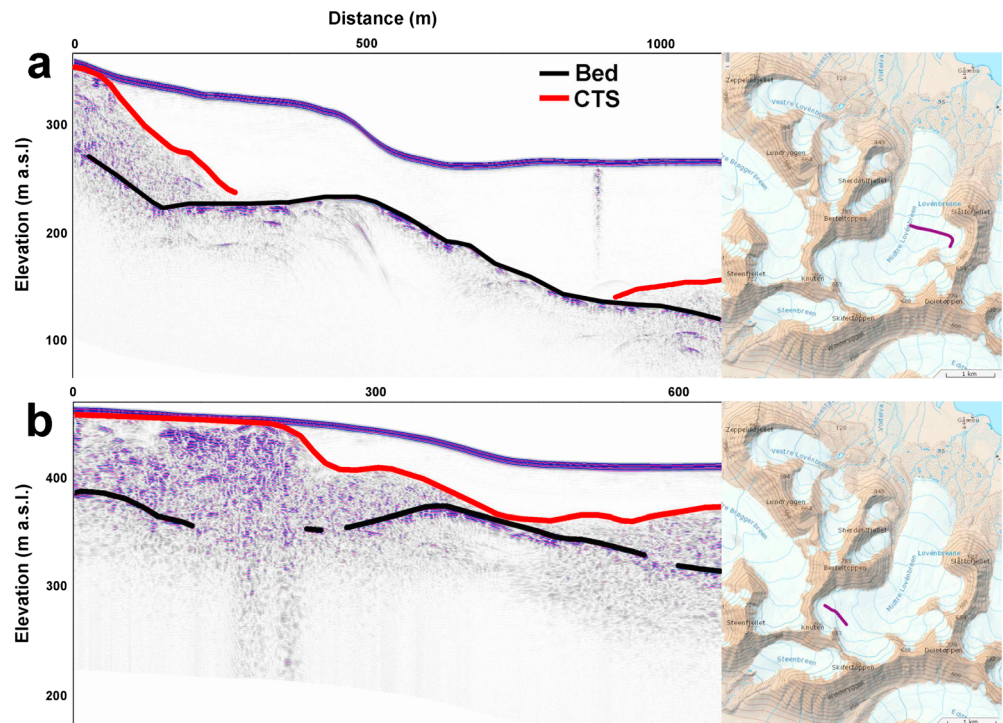
$$D = \sqrt{2Z\lambda + \frac{\lambda^2}{4}} \quad (1)$$

with  $Z$  for depth and  $\lambda$  for the antenna wavelength. At a depth of 100 m the horizontal resolution of our GPR system is equal to 18.3 m with the 100 MHz antenna, which doubles with the 25 MHz antenna.

Processing uncertainties encompass picking, migration, time-to-depth conversion as well as interpolation. Picking is subjective, and its quality depends also on the sharpness of the bed reflectors (as reported by Gusmeroli *et al.* [2012] and Pettersson *et al.* [2003]). Areas with multiple bed reflections that made picking challenging were not picked. We estimate picking error as 1–5 m. Crossover analyses were performed on each glacier to assess offsets between the bed reflector of valley-parallel and valley-transverse lines, as well as the migration quality (Table 4). Offsets were greatest close to the valley sides. For this reason we did not use valley-parallel lines close to the valley sides, only one centerline profile per glacier. Interpolation of the bed reflector was also assessed by measuring offsets between the interpolated surfaces and transverse profiles.

Time conversion to depth was performed using a constant radio wave velocity of  $167 \text{ m } \mu\text{s}^{-1}$  for all lines, regardless of their thermal divisions. Warm ice is characterized by higher propagation velocities than that of cold ice ( $140\text{--}160 \text{ m } \mu\text{s}^{-1}$  compared to  $167\text{--}170 \text{ m } \mu\text{s}^{-1}$ , respectively). For our largest glaciers this could mean a reduction in thickness less than 10% of their total thickness.

Finally positioning errors have to be accounted for. Four of our glaciers were surveyed with a single-frequency Garmin eTrex Legend. Although this GPS has a theoretical accuracy of 15 m, large “steps” in the vertical positioning could be observed on some of our radargrams. The centerlines of two large glaciers (Von Postbreen, Tunabreen) and the whole of Longyearbreen were surveyed using a higher-accuracy GPS in Real Time Kinematic 30/30 mode reaching a theoretical accuracy of 30 cm both in the horizontal and in the vertical. The GPS data were manually checked for errors and processed in stand-alone mode.



**Figure 8.** Two unmigrated profiles reaching the accumulation zone of Midtre Lovénbreen. (a) Profile collected in the lowermost cirque showing discontinuity between the central temperate core and temperate ice produced in the cirque. (b) Profile collected in the upper reaches of the glacier showing narrow connection between the central temperate core and temperate ice produced in the accumulation zone. Topographic maps from the Norwegian Polar Institute.

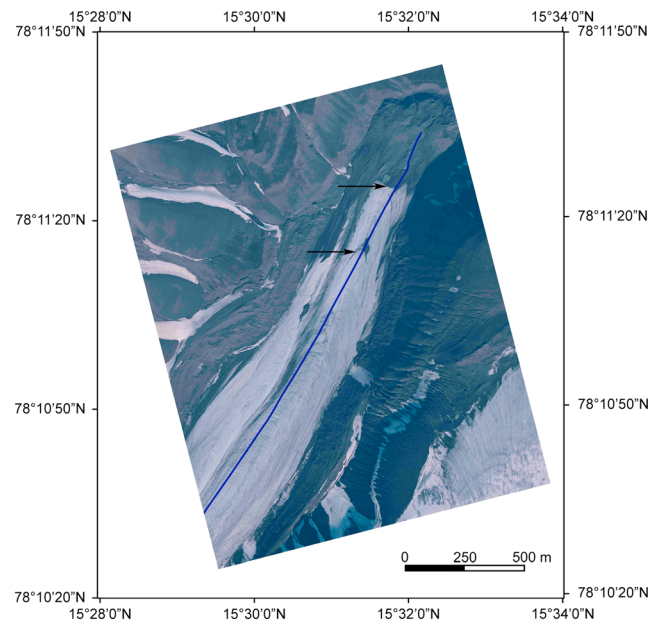
#### 4. Glacier Thermal Regimes

On Von Postbreen, a centerline profile from the front to the top of the accumulation zone shows a clear two-layered structure (see Figure 1d). In the ablation zone, the upper cold ice layer is 70–110 m thick and intersects the bed ~800 m from the terminus, indicating that the tongue of the glacier is entirely cold and frozen to the bed. Intermittent, narrow zones of cold ice also occur around the lateral margins of the glacier. Most of the glacier, however, is warm based, with temperate ice increasing in thickness to >200 m in the area of our survey. In the accumulation zone of Von Postbreen, more specifically in the tributary Potpeschniggbreen, the cold surface layer vanishes at an altitude of 725 m asl.

Kongsvegen has a similar two-layered structure, with an upper cold layer 50–95 m thick overlying temperate ice (Figure 2d). A frozen terminal zone is absent from Kongsvegen. The starting point of our survey line was located less than 500 m from the glacier front, but the calving cliff had retreated to this point by spring 2014. At that date, therefore, the glacier terminus consisted of a cold layer 60–70 m thick overlying a 30–40 m layer of warm ice. The glacier has narrow zones of cold ice around its lateral margins but is otherwise entirely warm based. Since the survey of Björnsson *et al.* [1996] in 1990, the point at which temperate ice intersects the surface has moved at least 2.5 km upglacier, and at the location of borehole K1 the CTS has been lowered by about 45 m.

The thermal structure of Tunabreen 12 years after its last surge is shown in Figure 3d. We can observe a thick layer of temperate ice from the beginning of our survey line (today approximately 1.7 km from the glacier front) all the way to 12 km upglacier. A layer of cold ice overlies the warm ice and varies in thickness from 100 to 25 m and appears to get thinner with elevation. Since the Bamber [1987] survey, the CTS has migrated closer to the surface by an average of 97 m. This centerline was surveyed both by a 100 MHz and 25 MHz antenna, and the depth to scattering is very similar between the two antennae.

Midtre Lovénbreen is not as thick as Von Postbreen or Kongsvegen, with maximum ice thicknesses of 170–180 m. The centerline profile shows a basal temperate layer 20 to 60 m thick, overlain by cold ice (Figure 4d). Our data do not extend to the glacier terminus due to deep snow at the time of our survey,



**Figure 9.** Aerial view of the lower 1.5 km of Longyearbreen (source: Norwegian Polar Institute 2011). The location of our survey line is indicated in blue. Arrows point to the englacial reflectors discussed in section 5.

but our surveys show wide fringes of cold ice around the lateral margins of the glacier. A survey line collected in the lowermost cirque shows temperate ice extending from the bed to the ice surface, although this is not connected to the warm core in the central part of the glacier (Figure 8a). Only the uppermost survey line shows a narrow connection between temperate ice in the upper part of the glacier and the central core (Figure 8b).

Tellbreen and Longyearbreen have similar thermal regimes to one another. Figures 5 and 6 clearly show the predominance of transparent, cold ice over the whole depth of both glaciers, confirming the observations of *Bælum and Benn* [2011] and *Etzelmüller et al.* [2000]. A basal temperature of about  $-4^{\circ}\text{C}$  was measured below the margin of Longyearbreen in 2001–2003 [*Humlum et al.*, 2005] (Figure 6). Longyearbreen

appears slightly thicker than Tellbreen with a maximum depth of 110 m. The steep V-shaped bed topography of Longyearbreen is the cause of clear off-nadir reflections. This made the interpretation of the radargrams difficult. *Etzelmüller et al.* [2000] tentatively suggested that there might be temperate ice below the upper section of the glacier, but our survey did not confirm this. *Bælum and Benn* [2011] found some evidence suggesting that a thin basal layer of temperate ice may exist under the thickest part of Tellbreen, but although our radargrams were particularly clear in this case we could not find any temperate ice (Figure 5).

In summary, Von Postbreen can be characterized as a polythermal glacier of type E [*Pettersson*, 2004]. That is, it is mostly composed of temperate ice with the exception of a cold surface layer in the ablation zone and lower accumulation zone. Tunabreen and Kongsvegen lack cold-based tongues and are therefore variants of Type E polythermal glaciers, in which the cold-based tongue is removed by ablation processes.

These thermal structures can be understood in terms of five processes: (1) formation of warm ice in the accumulation zone by latent heat released by refreezing of summer meltwater, (2) conductive cooling of near-surface ice due to heat losses to the atmosphere, (3) removal of cold ice in the ablation zone by melting (and calving in the case of Kongsvegen and Tunabreen), (4) strain heating by ice creep and basal motion, and (5) advection of heat through the system by ice flow. Calving is a particularly effective process in removing cold ice from glacier termini. Published work provides several additional examples of temperate ice layers extending to the termini of tidewater glaciers, including Negribreen [*Dowdeswell et al.*, 1984], Hansbreen [*Navarro et al.*, 2014], and Borebreen [*Bamber*, 1987]. Strain heating and ice advection during the 2003 surge can explain the observed thickening of the temperate layer on Tunabreen between 1983 and 2015. These processes are likely ineffective on the quiescent Von Postbreen and Kongsvegen, where the thermal structure appears to reflect the formation of warm firn in the accumulation zones, heat loss to the atmosphere where there is little or no net snow accumulation, and the removal of ice by ablation. In the case of Kongsvegen, we interpret thickening and expansion of the cold surface zone as the result of ongoing conductive cooling following termination of the 1948 surge.

Midtre Lovénbreen appears to be a glacier of type E in the process of becoming of type A. Although some accumulation still occurs in the cirques and refreezing of meltwater produces warm ice by latent heating, this temperate ice is barely connected to the core of temperate ice in the center of the glacier.

Tellbreen and Longyearbreen are both currently entirely cold and very slowly moving. Due to strongly negative mass balance in recent years, these glaciers have effectively no remaining accumulation zones.

Consequently, warming of snow and firn by refreezing is no longer an effective process. The current heat budget of these glaciers, therefore, is likely dominated by conductive cooling. There is, however, evidence that these glaciers, together with Midtre Lovénbreen, were more active in the past and that warm ice was formerly more extensive. This is discussed in the following section.

## 5. Evidence for Former Thermal Regimes

As described in section 2, Midtre Lovénbreen and Tellbreen both present strong evidence of past thermal switching. Longyearbreen has been less studied, but investigation of the newly collected radargrams reveals a large number of elongated internal reflectors extending from the bed toward the surface in the lower half of the glacier (see arrows in Figure 6). Similar englacial bands have been found close to the margins of Storglaciären, Sweden [Moore *et al.*, 2013]. The features midway up the glacier dip upglacier at about 28°, while they become less steep downglacier and reach a dip of 12° close to the terminus. Some are concave-up, increasing in dip from 13° to 51° as they approach the surface.

Projecting the GPR profile on a recent aerial photograph of the glacier shows that the englacial structures intersect a series of arcuate debris bands cropping out in the lower 500 m of Longyearbreen (see Figure 9). In the field, the uppermost debris bands consist of coarse, angular debris of probable supraglacial origin. These bands lie parallel to well-defined convex downglacier, arcuate ice foliation structures and appear to have formed by deposition of debris in the snowpack. Farther downglacier, however, striated clasts are found on the glacier surface indicating subglacial provenance. Although this evidence shows that part of the glacier may have been warm based at the Little Ice Age maximum, there are no surface structures indicative of thrusting and fracturing, as seen on Midtre Lovénbreen and Tellbreen. The presence of in situ paleosol and vegetation at the glacier bed [Humlum *et al.*, 2005], near the glacier margin 2 km up from the current glacier front suggests that any wet-based ice must have been confined to only the middle, thickest section of the glacier, while the thinner margins stayed covered by cold ice only moving by ice creep.

## 6. Discussion

The evidence presented above indicates that the small glaciers in our sample are less dynamically active now than in the past, reflecting changes in their thermal regime. The thicker parts of Midtre Lovénbreen remain warm based, but ice is becoming cold around the margins as the ice thins. Tellbreen is more advanced in this process, and the glacier is now entirely (or almost entirely) cold, although warm-based ice was formerly extensive. Longyearbreen displays the fewest signs of a dynamic past but still bears evidence that parts of the glacier were once warm based and erosive. Thinning and retreat of these glaciers from their LIA maxima, therefore, has been accompanied by a transition from an active, warm-based state to an inactive, cold state.

It follows that these glaciers must have undergone a corresponding thermal transition in the opposite direction during ice buildup prior to the LIA maximum. Accumulation of snow on top of permafrost would have initially produced thin, cold glaciers. Development of a warm core in their ablation zones could only occur once the ice was sufficiently thick to provide insulation against winter cooling. The transition to warm-based conditions could then result in a step change in rates of ice flow, causing additional heating from friction and the possibility of further flow acceleration and advance. There is evidence that several small glaciers in Svalbard advanced rapidly to their LIA maxima, including Midtre Lovénbreen, Drønbreen, and Scott Turnerbreen [Garwood and Gregory, 1898; Hagen *et al.*, 1993; Hodgkins *et al.*, 1999; Liestøl, 1988]. Whether this behavior should be referred to as surges has been widely debated. For example, Liestøl [1988] argued that Midtre Lovénbreen is surge type, based on the observation of a steep frontal bulge between 1860 and 1880. Hansen [2003] agreed that the glacier surged during the Little Ice Age but argued that it has now lost its ability to do so. In contrast, Hambrey *et al.* [2005] disagreed that the LIA advance was a surge and argued that this was normal behavior for an arctic polythermal glacier. Jiskoot *et al.* [2000] and King *et al.* [2008] also argued against Midtre Lovénbreen being a surge-type glacier. This disagreement, however, is largely semantic. The key point is that the small glaciers in our sample—and probably many other small glaciers in Svalbard—underwent thermal transitions over the course of the LIA, which correspond closely to the sequence of events invoked in the thermal switch model of glacier surges [Fowler *et al.*, 2001]. In the case of Midtre Lovénbreen and Tellbreen, the thermal switch resulted in a dynamic transition (a surge or surge-like advance), whereas for Longyearbreen the thermal switch affected a much more limited area, with no

evidence of a dynamic transition. This may reflect geometric factors, as Longyearbreen is narrower and less branched than the other two glaciers.

It is also noteworthy that land-terminating glaciers in Svalbard have mostly not produced minor retreat moraines during the twentieth century, despite there being climatic oscillations over this period [Nordli *et al.*, 2014]. These glaciers, therefore, appear to have transitioned directly from a dynamically active, advance phase to inactivity and decline, consistent with surges or “surge-like” behavior.

While small glaciers such as Tellbreen have reverted to a cold state following their active phase, large surge-type glaciers have not returned to a cold-based thermal regime following surges. This is not only the case for Tunabreen with its short surge cycle but also for Kongsvegen 65 years after surge termination and Von Postbreen after ~140 years. Both Von Postbreen and Kongsvegen are deep into their apparent “quiescent phases,” yet temperate ice is still present over most of their beds. Land-terminating glaciers such as Von Postbreen and Bakaninbreen [Murray *et al.*, 1997] are frozen to the bed in their terminal zones where ice is thin, but rapid removal of ice by calving prevents the development of cold tongues on tidewater glaciers such as Kongsvegen and Tunabreen.

Thus, in Svalbard three types of glaciers are known to have undergone switches from slow to fast flow: (1) small glaciers that underwent thermal cycles from cold to warm to cold based over the course of the LIA climate cycle, (2) large terrestrial glaciers which remain warm based throughout the whole surge cycle and develop a cold terminus while thinning during quiescence, and (3) large tidewater glaciers that remain warm based throughout the surge cycle (excepting narrow cold zones at their lateral margins).

This demonstrates two important points. First, thermal switching can explain the initiation of surges or surge-like advances of small glaciers over the Little Ice Age climatic cycle, but it cannot explain the cyclic surging behavior of large glaciers. Second, the absence of a frozen terminal zone on tidewater glaciers shows that thermal cycling plays no part in glaciers such as Tunabreen. On land-terminating glaciers, the existence of a cold tongue can influence surge evolution, and there are well-documented switches from cold-based to warm-based conditions during the downglacier propagation of surge fronts [Fowler *et al.*, 2001; Murray *et al.*, 1997]. Surge initiation, however, typically occurs in thicker, warm-bedded regions and is therefore unrelated to thermal transitions [e.g., Sund *et al.*, 2014; Fowler *et al.*, 2001; King *et al.*, 2015; Murray *et al.*, 2000]. Tidewater glaciers in many parts of the world can undergo “speed-ups” due to changes in the force balance at the terminus [e.g., Howat *et al.*, 2005; O’Neel *et al.*, 2005]. It is possible that some surges of Svalbard glaciers reflect tidewater dynamics [e.g., Murray *et al.*, 2003], although one well-documented tidewater glacier surge was initiated in warm-based ice in a tributary cirque, from where it propagated downglacier and activated the calving front [Kristensen and Benn, 2012]. The characteristics of tidewater glacier surges in Svalbard will be considered further in a forthcoming paper.

We conclude that thermal switching can explain surge-like behavior of small glaciers in Svalbard, but not surges of large, warm-based glaciers. Does this indicate that more than one surge mechanism is needed to explain the range of surge-type glaciers in Svalbard? Or, as expressed by Frappe-Seneclauze and Clarke [2007], does it imply that thermal regime is “collateral to surging phenomena rather than essential”?

We believe that these questions can be answered by looking beyond the issue of thermal regime and considering the *enthalpy balance* of the glacier [Sevestre and Benn, 2015]. In a glaciological context, enthalpy is defined as the internal energy of the glacier system, a function of ice temperature and water content [Aschwanden *et al.*, 2012]. Enthalpy balance is a more general concept than thermal regime and replaces the binary concept of cold versus warm with a more nuanced consideration of the evolving heat and water content of the system. Enthalpy may enter or leave a glacier through exchanges at its upper and lower boundaries and evolve within the glacier by conversion of potential energy during ice flow. At the glacier surface, enthalpy gains and losses include turbulent and radiative exchanges with the atmosphere coupled with conductive heat flux through the ice. Melting of snow followed by water storage results in net enthalpy gain, and enthalpy can be exchanged within the glacier when stored water contacts cold ice and refreezes, raising the ice temperature through latent heat release. At the bed, exchanges include geothermal heat and runoff of subglacial meltwater. Importantly, the bed can also gain enthalpy by the expenditure of potential energy during ice creep or sliding.

We argue that processes of enthalpy production and dissipation play a key role in glacier surging. Because strain rates and sliding speeds are sensitively dependent on ice temperature and water content, any change

in enthalpy at or near the bed will produce a dynamic response. If enthalpy gains exceed losses, the glacier will tend to accelerate; if enthalpy losses exceed gains, the glacier will slow down. Positive feedbacks between enthalpy production and glacier velocity may occur, depending on the relationship between stress and strain (i.e., friction laws), and the changing efficiency of enthalpy sinks (e.g., subglacial drainage). If a glacier is to remain close to steady state, therefore, there must be a broad equality between rates of buildup of gravitational potential energy (snow accumulation), conversion of potential energy into enthalpy by ice flow (mass flux rate), and loss of enthalpy by the dissipation of heat and meltwater.

We now apply these concepts to explain the past and present dynamic behavior of large and small Svalbard glaciers within a single framework. Under the present climate, Midtre Lovénbreen, Tellbreen, and Longyearbreen are small, thin, and slow-moving glaciers. They have not been observed to exhibit any dynamic instabilities for more than 100 years. These glaciers can therefore be categorized as being nonsurge-type glaciers under the climatic conditions of the twentieth and 21st centuries [Hansen, 2003; Jiskoot *et al.*, 2000]. They accumulate very little or no mass and are low enthalpy producers. Indeed, all three appear to be undergoing net enthalpy loss by heat conduction (and possibly evacuation of subglacial meltwater in the case of Midtre Lovénbreen).

However, conditions during the Little Ice Age led to sustained mass accumulation and positive enthalpy balance. Mass accumulation reduced conductive heat losses from the glacier beds and increased their potential energy store, increasing driving stresses and enthalpy production at the bed. Net gains in enthalpy led to positive feedbacks, progressively warming the bed then increasing the volume of stored water. In some cases, this ultimately toppled the glaciers into flow regimes above balance velocities. It is not known whether this occurred on Longyearbreen, suggesting that a thermal transition need not inevitably lead to dynamic instability. In contrast, other small Svalbard glaciers may have experienced two dynamic cycles during the LIA [Sletten *et al.*, 2001]. For small glaciers undergoing surges or surge-like advances in the late nineteenth or early twentieth centuries recovery and buildup were prevented by climate change and switching to a prolonged period of negative enthalpy balance. Dowdeswell *et al.* [1995] invoked recent climate change as a reason for the diminution of the number of surges in some regions, while Eisen *et al.* [2001] and Striberger *et al.* [2011] have revealed the control that climate exerts on mass balance and periodicity of surge cycles.

Many large glaciers in Svalbard undergo cyclic surging. During surges, thermal transitions occur only in the thin (<100 m), terminal parts of land-terminating glaciers, and tidewater glaciers may remain entirely warm based throughout. Enthalpy cycling on these glaciers, therefore, does not involve freezing and thawing of large areas of the bed but is expressed as fluctuations in the production and evacuation of basal meltwater. The large glaciers in our sample are thick (>300 m), and all contain large areas of temperate ice at their bed. For such thick glaciers under present climatic conditions, conductive heat losses are insufficient to cause freezing at the bed. Interestingly, the scale length used for glacier thickness in the Fowler *et al.* [2001] model is ~100 m, similar to that of the small glaciers in our sample, indicating that the model results are consistent with our conclusion that thermal cycling in Svalbard was limited to thin glaciers.

Studies of elevation changes have revealed that Svalbard glaciers with accumulation basins located at high altitude still accumulate mass today [e.g. Nuth *et al.*, 2007]. The buildup of mass shows that these glaciers are flowing at rates below the balance velocity, and potential energy is building up within the system. Enthalpy gains at the bed (due to gradually increasing driving stresses and ice creep rates) will gradually increase meltwater production. If this cannot be fully discharged via the subglacial drainage system, pressure builds up in the system and triggers a surge [Fowler *et al.*, 2001]. The surge may propagate downglacier into regions of cold-based ice, bringing larger areas of the bed to the melting point. The surge terminates when the water is able to drain either through the bed or through fractures in the glacier terminus [Smith *et al.*, 2002].

Thus, we argue that enthalpy cycling can explain the full spectrum of surges and surge-like behavior of Svalbard glaciers. Enthalpy cycles are expressed as a meltwater production-evacuation cycle on warm-based glaciers, and a thermal plus meltwater production-evacuation cycle on cold-based glaciers [Sevestre and Benn, 2015]. Surges occurring as nonlinear responses to a climatic signal can be regarded as *forced dynamic instabilities*, dictated by mismatches between their mass and enthalpy balances during the climate cycle. Conversely, cyclic surges occurring *within* a climate cycle may be termed *unforced dynamic instabilities*, reflecting out-of-phase oscillations in mass flux and enthalpy.

Some combinations of climate, topography, and bedrock geology permit glaciers to find steady state velocities, which match the balance mass flux with enthalpy losses. This condition appears to be most easily satisfied in cold, dry environments (thin, low-flux glaciers and efficient conductive heat losses) and warm, humid environments (high mass flux and meltwater discharges). The great majority of surging glaciers (including those in Svalbard) occur in intermediate climatic environments, where neither heat conduction nor runoff can effectively dissipate enthalpy from many glaciers [Sevestre and Benn, 2015]. Small glaciers in Svalbard were moved into this critical state during the LIA and removed from it by twentieth century warming. Large glaciers remain within the envelope and continue their endless struggle to reconcile their mass flux and enthalpy balance requirements.

## 7. Conclusions

We investigated the thermal regime of six glaciers in Svalbard by the means of new and published GPR profiles. The thinnest glaciers of the study are presently entirely cold, frozen to their beds or contain a “fossil” core of temperate ice. However, investigations of glaciological and geomorphological features have revealed evidence of past short-lived fast flow of at least two of these glaciers. These glaciers underwent a thermal cycle during the LIA, similar to the thermal switch mechanism described by Fowler *et al.* [2001]. As a response to the transition to warm state, their flow accelerated and produced a single surge-like episode before returning to a cold state.

The largest glaciers of the study are well-known surge-type glaciers. Our results clearly display their polythermal structure, with extensive temperate ice at their beds and the presence of a cold surface layer in their ablation zones and lower accumulation zones. Land-terminating glaciers have frozen tongues, where the winter cold wave can penetrate thin ice. Tidewater glaciers, however, may lack frozen tongues and are commonly entirely warm based. Tidewater glaciers may remain fully warm based between surges, indicating that thermal switching is not implicated in surge initiation. A cold front can influence the development and propagation of a surge, but it is not a fundamental condition for surging in Svalbard.

Enthalpy cycling can explain the former surge-like behavior of small glaciers and the cyclic surges of large glaciers. For small glaciers, the cycle is expressed as a thermal switch, whereas for large glaciers it is expressed as a cycle of basal meltwater production and drainage. Small glaciers are inferred to have undergone a thermal and dynamic cycle during the LIA, representing a nonlinear response to climatic forcing. They thus appear to have “become” surge-type glaciers during the optimal climatic conditions of the LIA but reverted to inactive states in the warming climate of the nineteenth and 21st centuries. In contrast, large glaciers remain surge type under present climatic conditions.

These conclusions can resolve the long-standing disagreement about the number of surge-type glaciers in Svalbard. Under present climatic conditions, surging behavior is confined to large glaciers (on average 68 km<sup>2</sup>), accounting for 21% of all glaciers. During the Little Ice Age, surges or surge-like instabilities affected the much larger population of small glaciers. Only a small fraction of these have been investigated, but our results suggest that thermal switching may have been very widespread and that Liestøl's claim that 90% of Svalbard's glaciers were “surge type” may be realistic for the period prior to the early twentieth century.

Classical surge cycles are a part of a continuum of dynamic instabilities that are more or less coupled to environmental forcing. When climatic conditions allow, glaciers can undergo a single episode of fast flow before returning into a “normal” state. There thus appears to have been a wave of surge-like advances during the LIA maximum across Svalbard.

## References

- AGF212 (2014), Field report 2014—Snow and ice processes, *Rep.*, The Univ. Centre in Svalbard, Longyearbyen, Norway.
- Annan, A. P. (1992), *Ground Penetrating Radar—Workshop Notes*, 128 pp., Sensors and Software Inc., Mississauga, Ont., Canada.
- Annan, A. P. (1999), *Ground penetrating radar—Workshop notes, Rep.*, 1091 pp., Sensors and Software Inc., Mississauga, Ont., Canada.
- Aschwanden, A., E. Bueler, C. Khroulev, and H. Blatter (2012), An enthalpy formulation for glaciers and ice sheets, *J. Glaciol.*, 58(209), 441–457.
- Bælum, K., and D. I. Benn (2011), Thermal structure and drainage system of a small valley glacier (Tellbreen, Svalbard), investigated by ground penetrating radar, *Cryosphere*, 5, 139–149.
- Bamber, J. L. (1987), Internal reflecting horizons in Spitsbergen glaciers, *Ann. Glaciol.*, 9, 5–10.
- Björnsson, H., Y. Gjessing, S.-E. Hamran, J. O. Hagen, O. Liestøl, F. Pálsson, and B. Erlingsson (1996), The thermal regime of sub-polar glaciers mapped by multi-frequency radio-echo sounding, *J. Glaciol.*, 42(140), 23–32.
- Clarke, G. K. C. (1976), Thermal regulation of glacier surging, *J. Glaciol.*, 16(74), 231–250.

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- Clarke, G. K. C., and R. H. Goodman (1975), Radio echo soundings and ice-temperature measurements in a surge-type glacier, *J. Glaciol.*, *14*(70), 71–78.
- Clarke, G. K. C., S. G. Collins, and D. E. Thompson (1984), Flow, thermal structure, and subglacial conditions of a surge-type glacier, *Can. J. Earth Sci.*, *21*(2), 232–240.
- Copland, L., and M. Sharp (2000), Radio-echo sounding determination of polythermal glacier hydrology, paper presented at Eighth International Conference on Ground Penetrating Radar.
- De Geer, G. (1910), Guide de l'excursion au Spitzberg, Stockholm, paper presented at Xle Congres Geologique International.
- Dowdeswell, J. A., D. J. Drewry, O. Liestøl, and O. Orheim (1984), Air-borne radio echo sounding of sub polar glaciers in Spitsbergen, in *Norsk Polarinstitutt Skrifter*, vol. 182, edited by J. A. Dowdeswell et al., 41 pp., Norsk Polarinstitutt, Oslo.
- Dowdeswell, J. A., R. Hodgkins, A.-M. Nuttall, J. O. Hagen, and G. S. Hamilton (1995), Mass balance change as a control on the frequency and occurrence of glacier surges in Svalbard, Norwegian High Arctic, *Geophys. Res. Lett.*, *22*(21), 2909–2912, doi:10.1029/95GL02821.
- Drewry, D. J., and O. Liestøl (1985), Glaciological investigations of surging ice caps in nordaustlandet Svalbard, *Polar Rec.*, *22*(139), 359–378.
- Eisen, O., W. D. Harrison, and C. F. Raymond (2001), The surges of Variegated Glacier, Alaska, U.S.A., and their connection to climate and mass balance, *J. Glaciol.*, *47*(158), 351–358.
- Etzelmüller, B., R. S. Ødegard, G. Vatne, R. S. Mysterud, T. Tønning, and J. L. Sollid (2000), Glacier characteristics and sediment transfer system of Longyearbreen and Larsbreen, western Spitsbergen, *Nor. Geogr. Tidsskr.- Norw. J. Geogr.*, *54*, 157–168.
- Fleming, E. J., H. Lovell, C. T. E. Stevenson, M. S. Petronis, D. Benn, M. J. Hambrey, and I. J. Fairchild (2013), Magnetic fabrics in the basal ice of a surge-type glacier, *J. Geophys. Res. Earth Surf.*, *118*, 2263–2278, doi:10.1002/jgrf.20144.
- Flink, A. E. (2013), Dynamics of surging tidewater glaciers in Tempelfjorden Spitsbergen, MSc thesis, 54 pp., Stockholm Univ., Stockholm, Sweden.
- Flink, A. E., R. Noormets, N. Kirchner, D. I. Benn, A. Luckman, and H. Lovell (2015), The evolution of a submarine landform record following recent and multiple surges of Tunabreen glacier, Svalbard, *Quat. Sci. Rev.*, *108*, 37–50.
- Forwick, M., T. O. Vorren, M. Hald, S. Korsun, Y. Roh, C. Vogt, and K. C. Yoo (2010), *Spatial and Temporal Influence of Glaciers and Rivers on the Sedimentary Environment in Sassenfjorden and Tempelfjorden, Spitsbergen*, *Geol. Soc. London Spec. Publ.*, *344*, 163–183.
- Fowler, A. C., T. Murray, and F. S. L. Ng (2001), Thermally controlled glacier surging, *J. Glaciol.*, *47*(159), 527–538.
- Fowler, C. M. R. (1990), *The Solid Earth: An Introduction to Global Geophysics*, Cambridge Univ. Press, Cambridge, U. K.
- Frappe-Seneclauze, T. P., and G. K. C. Clarke (2007), Slow surge of Trapridge glacier, Yukon Territory, Canada, *J. Geophys. Res.*, *112*, F03S32, doi:10.1029/2006JF000607.
- Garwood, E. J., and J. W. Gregory (1898), Contributions to the glacial geology of Spitsbergen, *Q. J. Geol. Soc. London*, *128*(4), 361–393.
- Glasser, N. F., and M. J. Hambrey (2001), Styles of sedimentation beneath Svalbard valley glaciers under changing dynamic and thermal regimes, *J. Geol. Soc.*, *158*, 697–707.
- Gusmeroli, A., P. Jansson, R. Pettersson, and T. Murray (2012), Twenty years of cold surface layer thinning at Storglaciären, sub-Arctic Sweden, 1989–2009, *J. Glaciol.*, *58*(207), 3–10.
- Hagen, J. O., O. Liestøl, E. Roland, and T. Jorgensen (1993), *Glacier Atlas of Svalbard and Jan Mayen, Meddelelser*, vol. 129, 141 pp., Norsk Polarinstitutt, Oslo.
- Hagen, J. O., K. Melvold, F. Pinglot, and J. A. Dowdeswell (2003), On the net mass balance of the glaciers and ice caps in Svalbard, Norwegian Arctic, *Arct. Antarct. Alp. Res.*, *35*(2), 264–270.
- Hambrey, A. (1894), En resa til norra Ishafet sommaren 1892, *Ymer*, *14*, 25–61.
- Hambrey, M. J., T. Murray, N. F. Glasser, A. Hubbard, B. Hubbard, G. Stuart, S. Hansen, and J. Kohler (2005), Structure and changing dynamics of a polythermal valley glacier on a centennial timescale: Midre Lovénbreen, Svalbard, *J. Geophys. Res.*, *110*, F01006, doi:10.1029/2004JF000128.
- Hansen, S. (2003), From surge-type to non-surge type glacier behaviour: Midre Lovenbreen, Svalbard, *Ann. Glaciol.*, *36*, 97–102.
- Hodgkins, R., and J. A. Dowdeswell (1994), Tectonic processes in Svalbard tide-water glacier surges: Evidence from structural glaciology, *J. Glaciol.*, *40*(136), 553–560.
- Hodgkins, R., J. O. Hagen, and S.-E. Hamran (1999), 20th century mass balance and thermal regime change at Scott Turnerbreen, Svalbard, *Ann. Glaciol.*, *28*, 216–220.
- Howat, I. M., I. Joughin, S. Tulaczyk, and S. Gogineni (2005), Rapid retreat and acceleration of Helheim glacier, East Greenland, *Geophys. Res. Lett.*, *32*, L22502, doi:10.1029/2005GL024737.
- Humlum, O., B. Elberling, A. Hormes, K. Fjordheim, O. H. Hansen, and J. Heinemeier (2005), Late-Holocene glacier growth in Svalbard documented by subglacial relict vegetation and living soil microbes, *Holocene*, *15*(3), 396–407.
- Jarvis, G. T., and G. K. C. Clarke (1975), The thermal regime of Trapridge glacier and its relevance to glacier surging, *J. Glaciol.*, *14*(71), 235–250.
- Jiskoot, H., T. Murray, and P. Boyle (2000), Controls on the distribution of surge-type glaciers in Svalbard, *J. Glaciol.*, *46*(154), 412–422.
- King, E. C., A. M. Smith, T. Murray, and G. W. Stuart (2008), Glacier-bed characteristics of midtre Lovénbreen, Svalbard, from high-resolution seismic and radar surveying, *J. Glaciol.*, *54*(184), 145–156.
- King, O., M. J. Hambrey, T. D. L. Irvine-Fynn, and T. O. Holt (2015), The structural, geometric and volumetric changes of a polythermal Arctic glacier during a surge cycle: Comfortlessbreen, Svalbard, *Earth Surf. Processes Landforms*, doi:10.1002/esp.3796.
- Kohler, J., T. D. James, T. Murray, C. Nuth, O. Brandt, N. E. Barrand, H. F. Aas, and A. Luckman (2007), Acceleration in thinning rate on western Svalbard glaciers, *Geophys. Res. Lett.*, *34*, L18502, doi:10.1029/2007GL030681.
- König, M., C. Nuth, J. Kohler, G. Moholdt, and R. Pettersen (2013), A digital glacier database for Svalbard, in *Global Land Ice Measurements From Space: Satellite Multispectral Imaging of Glaciers*, edited by J. S. Kargel et al., chap. 10, pp. 229–239, Praxis-Springer, Berlin.
- Kristensen, L., and D. I. Benn (2012), A surge of the glaciers Skobreen-Paulabreen, Svalbard, observed by photographs and remote sensing data, *Polar Res.*, *31*, 1–9, doi:10.3402/polar.v31i0.11106.
- Lefauconnier, B. (1987), Fluctuations glaciaires dans le Kongsfjord, 79°N, Spitsbergen, Svalbard; analyse et conséquences, PhD thesis, p. 252, Univ. de Grenoble, Grenoble, France.
- Lefauconnier, B., J. O. Hagen, and J. P. Rudant (1994), Flow speed and calving rate of Kongsbreen glacier, Svalbard, using SPOT images, *Polar Res.*, *13*(1), 59–65, doi:10.1111/j.1751-8369.1994.tb00437.x.
- Liestøl, O. (1988), The glaciers in the Kongsfjorden area, Spitsbergen, *Nor. Geogr. Tidsskr.- Norw. J. Geogr.*, *42*(4), 231–238.
- Lovell, H., E. J. Fleming, D. Benn, B. Hubbard, S. Lukas, and K. Naegeli (2015), Former dynamic behaviour of a cold-based valley glacier on Svalbard revealed by basal ice and structural glaciology investigations, *J. Glaciol.*, *61*(226), 309–328, doi:10.3189/2015JoG14J120.
- Mayer, C., A. C. Fowler, A. Lambrecht, and K. Scharer (2011), A surge of North Gasherbrum Glacier, Karakoram, China, *J. Glaciol.*, *57*(204), 904–916.
- Melvold, K., and J. O. Hagen (1998), Evolution of a surge-type glacier in its quiescent phase: Kongsvegen, Spitsbergen, 1964–95, *J. Glaciol.*, *44*(147), 394–404.

- Moore, P. L., N. R. Iverson, K. T. Uno, M. P. Dettinger, K. A. Brugger, and P. Jansson (2013), Entrainment and emplacement of englacial debris bands near the margin of Storglaciären, Sweden, *Boreas*, *42*, 71–83.
- Murray, T., D. L. Gooch, and G. W. Stuart (1997), Structures within the surge front at Bakaninbreen using ground penetrating radar, *Ann. Glaciol.*, *24*, 122–129.
- Murray, T., G. W. Stuart, P. J. Miller, J. Woodward, A. M. Smith, P. R. Porter, and H. Jiskoot (2000), Glacier surge propagation by thermal evolution at the bed, *J. Geophys. Res.*, *105*(B6), 13,491–13,507, doi:10.1029/2000JB900066.
- Murray, T., A. Luckman, T. Strozzi, and A. M. Nuttall (2003), The initiation of glacier surging at Fridtjovbreen, Svalbard, *Ann. Glaciol.*, *36*(1), 110–116.
- Navarro, F. J., and O. Eisen (2009), Ground-penetrating radar, in *Remote Sensing of Glaciers*, edited by P. Pilleka and W. G. Rees, pp. 195–229, Taylor & Francis, London.
- Navarro, F. J., A. Martín-Español, M. Grabiec, J. Otero, E. V. Vasilenko, and D. Puczo (2014), Ice volume estimates from ground-penetrating radar surveys, Wedel Jarlsberg Land Glaciers, Svalbard, *Arct. Antarct. Alp. Res.*, *46*(2), 394–406.
- Nordli, Ø., R. Przybylak, A. E. J. Ogilvie, and K. Isaksen (2014), Long-term temperature trends and variability on Spitsbergen: The extended Svalbard Airport temperature series, 1989–2012, *Polar Res.*, *33*, 21349, doi:10.3402/polar.v33.21349.
- Nuth, C., J. Kohler, H. F. Aas, O. Brandt, and J. O. Hagen (2007), Glacier geometry and elevation changes on Svalbard (1936–90): A baseline dataset, *Ann. Glaciol.*, *46*(1), 106–116.
- Nuth, C., G. Moholdt, J. O. Hagen, and A. Kääb (2010), Svalbard glacier elevation changes and contribution to sea level rise, *J. Geophys. Res.*, *115*, F01008, doi:10.1029/2008JF001223.
- Ødegard, R. S., J. O. Hagen, and S.-E. Hamran (1997), Comparison of radio-echo sounding (30–1000 MHz) and high-resolution borehole-temperature measurements at Finsterwalderbreen, southern Spitsbergen, Svalbard, *Ann. Glaciol.*, *24*, 262–267.
- Oélsner, C. (1966), Ergebnisse von Gravimeternmessungen im Kingsbay-Gbeite (Westspitsbergen), *Petermanns Geogr. Mitt.*, *110*(2), 111–116.
- O’Neel, S., W. T. Pfeffer, R. M. Krimmel, and M. F. Meier (2005), Evolving force balance at Columbia glacier, Alaska, during its rapid retreat, *J. Geophys. Res.*, *110*, F03012, doi:10.1029/2005JF000292.
- Paterson, W. S. B. (1994), *The Physics of Glaciers*, 3rd ed., 480 pp., Elsevier, Oxford, New York, and Tokyo.
- Pettersson, R. (2004), *Dynamics of the Cold Surface Layer of Polythermal Storglaciären*, Stockholm Univ., Sweden.
- Pettersson, R. (2005), Frequency dependence of scattering from the cold-temperate transition surface in a polythermal glacier, *Radio Sci.*, *40*, RS3007, doi:10.1029/2004RS003090.
- Pettersson, R., P. Jansson, and P. Holmlund (2003), Cold surface layer thinning on Storglaciären, Sweden, observed by repeated ground penetrating radar surveys, *J. Geophys. Res.*, *108*(F1), 6004, doi:10.1029/2003JF000024.
- Plassen, L., T. O. Vorren, and M. Forwick (2004), Integrated acoustic and coring investigation of glacial deposits in Spitsbergen fjords, *Polar Res.*, *23*, 89–110.
- Rippin, D., I. Willis, and A. Neil (2005), Seasonal patterns of velocity and strain across the tongue of the polythermal glacier midre Lovénbreen, Svalbard, *Ann. Glaciol.*, *42*(1), 445–453.
- Robin, G. D. Q. (1955), Ice movement and temperature distribution in glaciers and ice sheets, *J. Glaciol.*, *2*(18), 523–532.
- Robin, G. D. Q. (1969), Initiation of glacier surges, *Can. J. Earth Sci.*, *6*(4), 919–928.
- Sevestre, H., and D. Benn (2015), Climatic and topographic controls on the global distribution of surge-type glaciers: Implications for a unifying model of surging, *J. Glaciol.*, doi:10.3189/2015JoG14J136.
- Sletten, K., A. Lyså, and I. Lønne (2001), Formation and disintegration of a high-arctic ice-cored moraine complex, Scott Turnerbreen, Svalbard, *Boreas*, *30*(4), 272–284.
- Smith, A. M., T. Murray, B. M. Davison, A. F. Clough, J. Woodward, and H. Jiskoot (2002), Late surge glacial conditions on Bakaninbreen, Svalbard, and implications for surge termination, *J. Geophys. Res.*, *107*(B8), 2152, doi:10.1029/2001JB000475.
- Striberger, J., S. Jbörck, I. Ö. Benediktsson, I. Snowball, C. B. Uvo, O. Ingólfsson, and K. H. Kjær (2011), Climatic control on the surge periodicity of an Icelandic outlet glacier, *J. Quat. Sci.*, *26*(6), 561–565.
- Sund, M., T. Eiken, J. O. Hagen, and A. Kaab (2009), Svalbard surge dynamics derived from geometric changes, *Ann. Glaciol.*, *50*(52), 50–60.
- Sund, M., T. R. Lauknes, and T. Eiken (2014), Surge dynamics in the Nathorstbreen glacier system, Svalbard, *Cryosphere*, *8*, 623–638.
- Voigt, U. (1965), Die Bewegung der Gletscherzunge des Kongsvegen (Kingsbay, Westspitsbergen), *Petermanns Geogr. Mitt.*, *109*(1), 1–8.
- Woodward, J., T. Murray, and A. McCaig (2002), Formation and reorientation of structure in the surge-type glacier Kongsvegen, Svalbard, *J. Quat. Sci.*, *17*(3), 201–209.